PB89-206650

Surface Inhomogeneity Effects on Convective Diffusion

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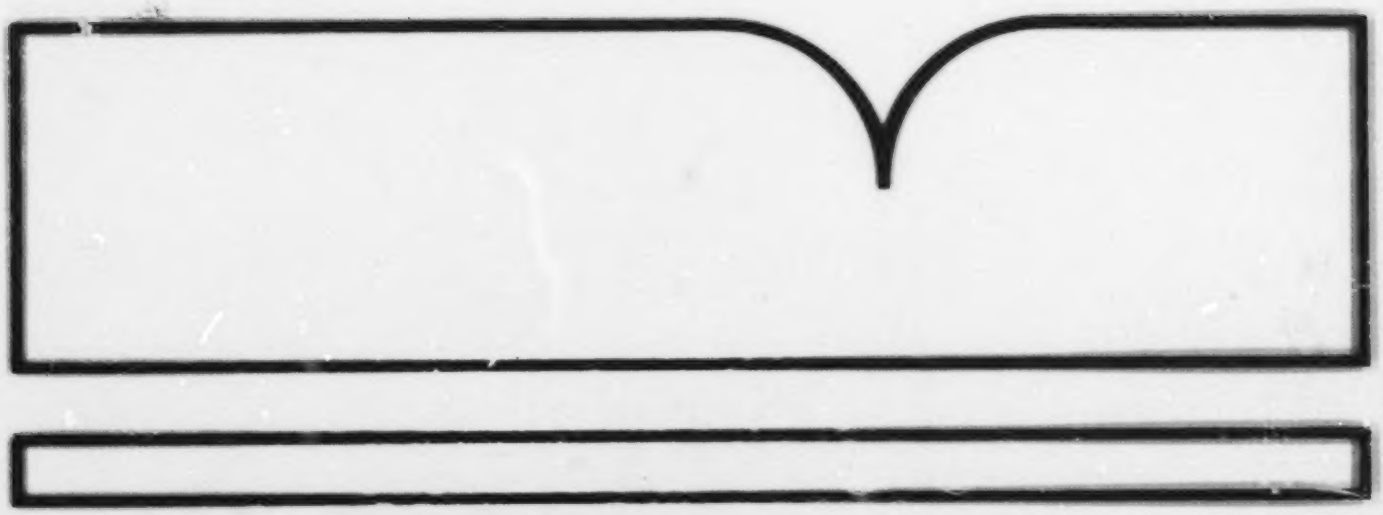

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PB89-206650

1. REPORT NO.	2.	3. F
EPA/600/J-88/334		

4. TITLE AND SUBTITLE
Surface Inhomogeneity Effects on Convective Diffusion

5. REPORT DATE

6. PERFORMING ORGANIZATION CODE

7. AUTHOR(S)
Gary A. Briggs

8. PERFORMING ORGANIZATION REPORT NO.

9. PERFORMING ORGANIZATION NAME AND ADDRESS
Same block 12.

10. PROGRAM ELEMENT NO.
A119/A/01/03-6084 FY-88)

11. CONTRACT/GRANT NO.

12. SPONSORING AGENCY NAME AND ADDRESS
Atmospheric Sciences Research Laboratory - RTP, NC
Office of Research and Development
Environmental Protection Agency
Research Triangle Park, NC 27711

13. TYPE OF REPORT AND PERIOD COVERED

14. SPONSORING AGENCY CODE
EPA/600/09

15. SUPPLEMENTARY NOTES
Boundary-Layer Meteorology 45 (1988) 117-135

16. ABSTRACT

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17. KEY WORDS AND DOCUMENT ANALYSIS

a. DESCRIPTORS	b. IDENTIFIERS/OPEN ENDED TERMS	c. COSATI Field/Group

REPRODUCED BY
U.S. DEPARTMENT OF COMMERCE
NATIONAL TECHNICAL INFORMATION SERVICE
SPRINGFIELD, VA. 22161

18. DISTRIBUTION STATEMENT	19. SECURITY CLASS *(This Report)*	21. NO. OF PAGES
RELEASE TO PUBLIC	UNCLASSIFIED	21
	20. SECURITY CLASS *(This page)* UNCLASSIFIED	22. PRICE A01/3.95 $6.95

EPA/600/J-88/334

SURFACE INHOMOGENEITY EFFECTS ON CONVECTIVE DIFFUSION

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(Received in final form 5 May, 1988)

Abstract. It is suggested that convective scaling, with appropriate extensions, provides the most useful framework for estimating the effects of urban-scale surface inhomogeneities on diffusion in convective conditions. Strong contrasts in surface heat flux exist between cropland, forests, urban areas, and water or marshland surfaces. It is argued that a typical fetch for convective turbulence to readjust to changed heat (or buoyancy) input from the surface below is $2(U/w^*)h$, where U is the mean wind speed in the mixing layer, w^* is the convective scaling velocity, and h is the mixing depth. In contrast, the fetch required for wind speed to readjust to new underlying surface roughness is of the order $(U/u^*)^2 h/2$, where u^* is the friction velocity.

The ratio w^*/U is the best index of diffusion rates in moderately to very unstable conditions. General urban effects on heat flux, h, and U are discussed separately, then their combined effects on w^*/U are estimated. While this ratio can double over a large city during light winds, its increase is much less for small cities, or during moderate winds. Finally, some examples of heat flux inhomogeneities causing stationary convective features are presented. Steady downdrafts associated with these features are of the order of $0.4w^*$, and could significantly increase surface concentrations from elevated sources.

1. Introduction**

This paper discusses, in semi-quantitative fashion, some important physical phenomena relevant to modeling short-range diffusion subject to effects of surface inhomogeneities in convective conditions. "Short-range" in this context means scales of the order of the width of a large urban area, say 30 km. The scales of surface inhomogeneities discussed extend upward from a fraction of a kilometer, or a few tenths of the mixing depth. These inhomogeneities can be due to rural-urban, forest-field, land-water or other boundaries between surfaces with large differences in sensible heat fluxes. Transport and fumigation caused by sea or lake breezes are not discussed here. "Convective conditions" normally are present over land during the day, except within an hour of sunrise or sunset, during high winds, under overcast skies, or over snow or fog cover. More specifically, these conditions correspond to $h/|L| > 10$, where h is the mixing depth and L is the Obukhov length; in such situations, buoyancy is the dominant

* On assignment from the National Oceanic and Atmospheric Administration, U.S. Department of Commerce.

** This paper is based on a presentation made at the AMS Specialty Conference on Air Quality Modeling of the Urban Boundary Layer, in Baltimore, late 1983.

Boundary-Layer Meteorology **45** (1988) 117–135.
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force that drives turbulence through the mixing layer except for vertical turbulent motions very close to the surface. The applicability of the results extends to a variety of site situations near significant surface inhomogeneities, not solely those due to urbanization.

2. Variables Characterizing Convective Diffusion

For the past several decades, two methods have been widely used for classifying atmospheric stability to order diffusion data and make diffusion predictions. Pasquill (1961) proposed wind speed and insolation or nighttime cloudiness as classifiers, with wind speed having the stronger effect. Briggs (1982) showed that, in the daytime, wind speed cubed divided by insolation strongly correlates with L. This length has been shown to order vertical diffusion near the ground quite effectively, but it provides cruder ordering of lateral diffusion rates (Briggs, 1988). The "Brookhaven method" (Smith, 1973) uses a single stability index, the variance of horizontal wind direction, σ_a. It is not surprising that σ_a correlates well with lateral dispersion, σ_y, because the "waved garden hose" analogy and statistical theory imply that $\sigma_y \approx \sigma_a x$ near the source (x = distance downwind of the source). More surprising is the observation that in the daytime and near the source, σ_a also correlates well with vertical dispersion, whereas at nighttime this correlation is nil (Cramer *et al.*, 1958). A third classifying method, using temperature gradient alone, was shown to be essentially uncorrelated with daytime diffusion rates (Weber *et al.*, 1977; Briggs and McDonald, 1978). Although it correlates strongly with surface heat flux, it has little correlation with daytime wind speed, which is the stronger factor affecting diffusion.

However, convective diffusion is most vigorously carried out by large eddies limited vertically by h, the average height of the well-mixed layer. A major shortcoming of the above methods is that they do not directly account for the effect of h, which becomes the most important meteorological variable affecting concentrations as distance increases. It is unimportant only for vertical diffusion, from very low sources, at short distances (less than 1 km).

An entirely new approach to modeling daytime diffusion has grown out of scaling ideas proposed by Deardorff (1970a), who was inspired by Busse's (1970) scaling of convective turbulent motions generated by a constant heat flux between two horizontal plates. For atmospheric boundary layers dominated by convectively driven turbulence, Deardorff suggested that: (1) large eddy motions scale to h, and (2) turbulent velocities are proportional to the convective velocity scale, $w^* = (H^* h)^{1/3}$. Here, H^* is approximately $(g/c_p\rho\overline{T})$ times the surface sensible heat flux, which is equivalent to $(g/T)\overline{w'T'}$ (g is gravitational acceleration, c_p is the specific heat capacity of air, ρ is the density of air, T is the absolute temperature of air, and w is the local vertical velocity; bars denote time means, and primes denote deviations from the mean). Over moist surfaces, a more exact definition of H^* should be used, with virtual absolute temperature in

place of T in the last expression. This is the vertical flux of buoyant accelerations produced by virtual temperature fluctuations; alternatively H^* can be understood as flux of buoyant force divided by the density of the medium. A simple way to understand the physical basis of w^* is to consider the rise of a single thermal with a vertical velocity $\bar{w}$ that carries an acceleration flux proportional to H^*. The buoyant acceleration is then $\sim H^*/\bar{w}$, and assuming that the kinetic energy is proportional to the potential energy released as the thermal traverses h, we have $\bar{w}^2 \sim (H^*/\bar{w})h$. This gives $\bar{w} \sim w^*$.

Deardorff (1970b) first showed that convective scaling brought aircraft measurements of turbulence into good agreement with predictions of a large-eddy numerical simulation. The ability of this $w^* - h$ scaling to order field measurements of convective turbulence was soon demonstrated in the Minnesota 1973 experiments and many others (Kaimal *et al.*, 1976; Caughey and Palmer, 1979; Druilhet *et al.*, 1983; Hicks, 1985).

Because atmospheric diffusion results from turbulence, a scaling that orders turbulence data should do the same for diffusion data. This notion was verified for convective diffusion in a well known series of laboratory experiments carried out in a water tank with a heated bottom (Willis and Deardorff, 1974, 1976, 1978, 1981). These results were supported by results of large-eddy numerical simulations (Lamb, 1978, 1979). Vertical diffusion was observed to depart markedly from predictions of customary Gaussian plume models. Plumes from surface releases tended to lift off the surface in the tank after a time $\approx 0.5h/w^*$, followed by lofting of most of the material into the upper half of the mixing layer, while maximum concentrations from elevated releases tended to descend at a rate $\approx 0.5w^*$ until impacting the surface, producing higher surface concentrations than conventionally modeled.

For relatively homogeneous terrain, there is now a substantial amount of diffusion field data that essentially confirms the results of the laboratory and numerical simulations, when the data are nondimensionalized in terms of convective scaling (height by h, distance by hU/w^*, and concentration by $Q/(Uh^2)$, where Q is release rate and U is the mean wind speed in the mixing layer) (Nieuwstadt, 1980; Moninger *et al.*, 1983; Eberhard *et al.*, 1985; Briggs, 1985; Briggs *et al.*, 1986). A more detailed account of the historical development of convective scaling and its applications to diffusion over homogeneous terrain is given in Briggs (1985).

In the following, we consider some implications of convective scaling for diffusion in convective conditions over inhomogeneous surfaces, extending some of the elementary physical ideas behind this scaling to inhomogeneous situations. Specifically, we shall examine: (1) the rate of adjustment of convective turbulence velocities to large increases or decreases in H^*, (2) the impact on diffusion rates of typical changes in H^*, h, and U over surface anomalies like cities, and (3) the intensity of stationary convective rolls caused by hot or cold spots such as industrial parks or water surfaces.

3. Response of Turbulence to Heat Flux Inhomogeneities

Flights over St. Louis with turbulence-instrumented aircraft, reported by Ching (1985), showed that turbulent fluxes at 150 m are highly inhomogeneous. On one day, measured heat fluxes averaged over flight segments 20 km ranged from 20 W m^{-2} over the Mississippi River to six times as large over the downwind side of the city (Figure 1). What kind of fetch over large changes in H^* is required for convective turbulence to respond to changed w^*? It will be argued that the response distance for turbulence energy readjustments is on the order of several $(U/w^*)h$, typically several km. Because this is the convective scaling distance, it also is the approximate distance of travel needed for complete vertical mixing. This applies to the mixing of horizontal momentum as well as concentration. Thus, several $(U/w^*)h$ is sufficient distance for wind shear and u^* to adjust substantially to a changed surface in convective conditions. However, we shall see that a much longer distance is required to obtain a new equilibrium value of U.

Suppose that a larger heat flux is encountered, so w^* is increased over its initial value, w_0^*. Turbulence velocities do not respond immediately, because the increased heat flux must rise all the way through h to release fully its potential energy, which is the source of convective turbulent energy. From Willis and Deardorff (1976, Figure 2), we see that the rise rate of the mean particle height, $\bar{z}$, is about $0.5w^*$ for a surface release. Because heat flux also arises from the surface, and the upward velocities at first scale to w_0^*, it takes approximately $h/(0.5w_0^*)$ time for the new heat flux to rise through the whole layer. The travel distance for the new heat flux to be fully effective is this time multiplied by U, about $2(U/w_0^*)h$.

What happens if heat flux is suddenly absent, as when air passes over a cool body of water, for instance? Turbulence energy in a steady convective boundary layer dissipates at a rate about $0.5H^*$, balancing the mean rate of production of such energy. This dissipation rate can also be expressed in terms of $\sigma^2 = \sigma_{u^2} + \sigma_{v^2} + \sigma_{w^2}$, the total velocity variance. Each component of σ averages approximately $0.6w^*$ through the mixing layer (Briggs, 1985), so $\sigma^3 \approx 1.1w^{*3} = 1.1H^*h$ and turbulent kinetic energy dissipation rate $\approx 0.5H^* \approx 0.45\sigma^3/h$ in the case of steady H^*. When the driving force H^* is removed at $t = 0$, let us assume that $\mathrm{d}(\sigma^2/2)/\mathrm{d}t = -0.45\sigma^3/h$, where $\sigma^2/2$ is the total specific kinetic energy of turbulence. This equation has a simple solution:

$$\sigma/\sigma_0 = (1 + 0.45\sigma_0 t/h)^{-1}. \tag{1}$$

The velocity standard deviation reduces to half of its initial value at a time $t = 2.2h/\sigma_0 = 2.1h/w_0^*$, or a distance $x \approx 2(U/w_0^*)h$. (A very similar equation was developed by Brost and Nieuwstadt (1985); they compared it to results of large-eddy numerical simulations; these simulations showed 50% reduction in σ at tw_0^*/h about 10% larger than the above estimate.) Thus, $X \approx 2$ seems to be

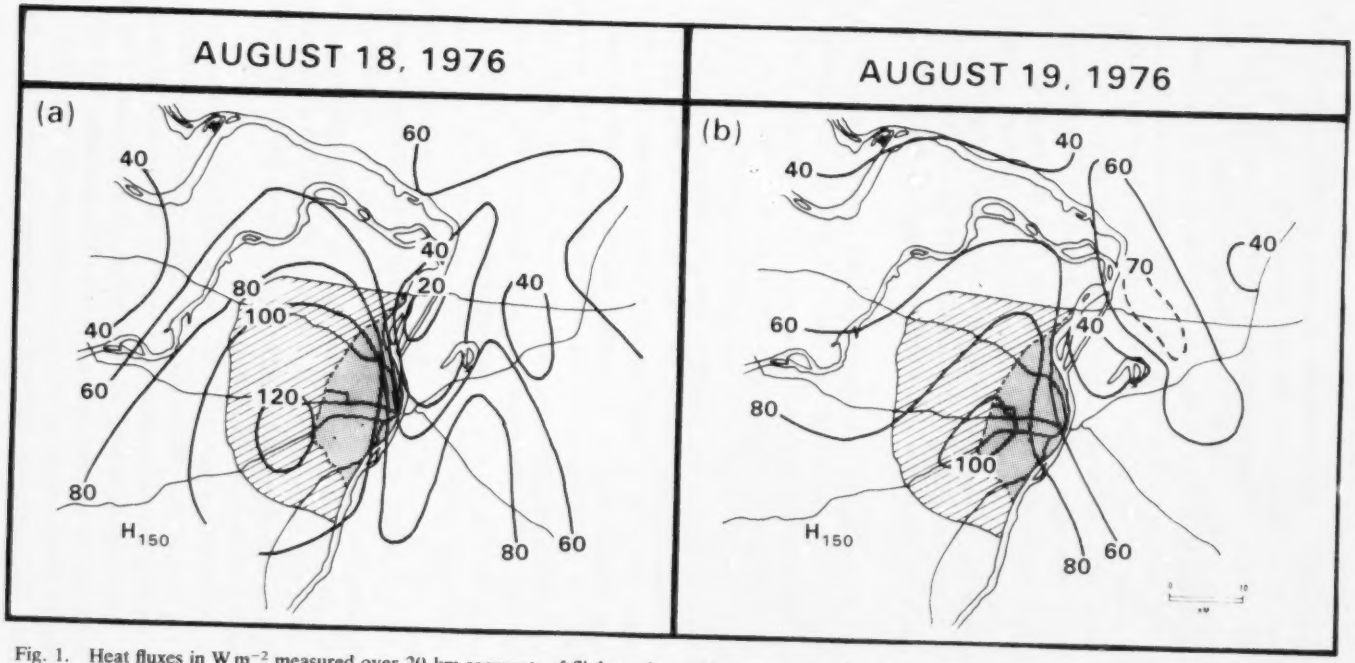

Fig. 1. Heat fluxes in W m^{-2} measured over 20-km segments of flight paths ~150 m AGL over the St. Louis area near midday (Ching, 1985). Winds were from the east at 3.1 m s^{-1} on the 18th and 3.5 m s^{-1} on the 19th.

generally the fetch required for substantial response to changes in heat flux, where $X = (x/U)w^*/h$ is dimensionless distance in convective scaling.

How well do the velocity variances measured in the St. Louis RAPS program (Ching, 1985) respond to changes in w^*? Any change in h over the city must be neglected because h was measured in only one location; h was about 1000 m on 18 August 1986 and 1500 m the following day. In any case, percentage changes in H^* were very much larger. Winds were from the east at $\approx 3\ \mathrm{m\ s^{-1}}$. On both days, the local values of σ_u/w^* and of σ_w/w^* measured by aircraft at $z = 150$ m were within $\pm 10\%$ of their mean values over approximately 80% of the area. The mean values of these ratios depend slightly on how heat flux measurements are extrapolated to the surface. If H^* is taken to be its surface value times $(1 - 2.4z/h)$ below $0.2h$, an assumption supportable by Caughey and Palmer (1979, Figure 3), then σ_u/w^* averaged 0.63 and 0.65 on the two days, and σ_w/w^* averaged 0.62 and 0.60. These are well within the ranges of other experimental values.

The overflight values of σ_w/w^* remained very close to 0.6 over broad areas of high heat flux (Figure 1) in and around the city on both days. The distance from the edge to the center of the high heat flux was ~20 km, which translates to $X \cong 10$ and 6 on these two days, using $w^* = w_0^*$. Thus, there was ample distance for σ_w to respond to the new values of w^*, 20 to 25% larger than w_0^*. The σ_w/w^* ratio deviated most in areas of low heat flux, where σ_w did not drop as much as w^*. The most extreme example was on the 18th, on a N–S flight leg centered over the Mississippi River at Mosentheim and Gaberet Islands, just NE of the city center. These islands are heavily wooded and largely undeveloped, being mostly less than 5 m above the river, but both mainland shores are heavily developed. This island area, including water surface, is about 4 km wide and 12 km long, oriented NNE. The measured heat flux over this area was only $20\ \mathrm{W\ m^{-2}}$, compared with $60\ \mathrm{W\ m^{-2}}$ measured upwind (see Figure 1). The observed σ_w showed no drop at all, but remained within 5% of upwind and downwind values. The same was true on the 19th, but the heat flux dropped only half as much. For the center of these islands, the fetch of the easterly wind from the developed shore was about 2 km. With $w_0^* \approx 1.37\ \mathrm{m\ s^{-1}}$, $U_0 \approx 3\ \mathrm{m\ s^{-1}}$, and $h = 1000$ m, this gives $X \approx 0.9$. Apparently, this was not enough fetch to affect σ_w measurably at $0.15h$ (some of the heat flux from the upwind shore would still be rising and driving turbulence at this point). The patterns for σ_u were similar to the σ_w patterns, reinforcing the above conclusion.

The best measurements of turbulence at a very high heat flux site were of σ_v on a tower 1 km west of the Mississippi River. This commercial-industrial site was largely surrounded by low buildings and blacktop, with almost no vegetation. On eight days with easterly winds, the heat fluxes measured here were 3 to 4 times those measured at an upwind, rural tower. This probably was caused by rapid heat transfer to air cooled by an overwater trajectory. The rural value of σ_v/w^* averaged 0.62, agreeing with other experimental values. At the com-

mercial-industrial site, however, σ_v increased by about 15%, while w^* increased by about 50%; thus, σ_v/w^* averaged only 0.47 at this site. This partial response to increased w^* seems in line with the $X \approx 2$ requirement for full response to H^* change, considering the modest dimensionless fetch from the Mississippi shoreline on these days, roughly $X \approx 1/2$.

4. Heat Flux Inhomogeneities in Urban Areas

Increased heat flux in urban areas has often been noted. Ching *et al.* (1983a), during their 1976 study of St. Louis, noted that average midday values of surface heat flux for the month of August were about 120 W m^{-2} at a rural site, 135 W m^{-2} in a residential area with many trees, and 295 W m^{-2} at a commercial-industrial site nearly void of vegetation. At midday, the ratio of sensible to latent heat fluxes, the Bowen ratio, was about 2.0 at the industrial site and 0.6 at the rural site, showing the important role of vegetation in distributing solar energy. The Bowen ratio can be as small as 0.2 over wet cropland, with much-reduced sensible heat flux; the heat flux could be four times as large in a relatively dry city. More typically, values of urban H^* are about twice rural values, as seen in Figure 1. The flux of latent heat also makes a small contribution to the buoyancy flux H^*, because water vapor is lighter than air. However, the ratio of this contribution to H^* compared to $(g/T)\overline{w'T'}$ is about 0.07 divided by the Bowen ratio. Because of the reduced effectiveness of latent heat in producing buoyancy, in heavily irrigated desert cities there may be a reversed urban effect; H^* may be less within the city, because a substantial fraction of insolation provides energy for evaporation of sprayed water and transpiration from vegetation.

5. Mixing Depth "Doming" Over Urban Areas

The haze layer sometimes forms a visible dome over a city, suggesting an enhanced mixing depth. Wong and Dirks (1978) reported a July day in St. Louis with a measured h about 40% larger in the city than at the rural rawinsonde station, and Spangler and Dirks (1974) presented airborne lidar measurements showing doming of h over the same city on August days, with about 25% increases in h. However, a systematic 30-month rawinsonde study of St. Louis showed much smaller rural-urban h differences (Schiermeier, F. A., personal communication). In fact, averages over all seasons showed no significant differences in daytime h values between the urban and the rural sites. For the two summers studied, however, the average urban h at noon was higher than the average rural value by 24% in 1975 and 5% in 1976. Thus, the average h enhancement may vary considerably.

One might suspect that such "doming" is due to lifting caused by longitudinal convergence as air entering a city slows down. However, an order-of-magnitude analysis shows that there is not enough kinetic energy available during con-

vective conditions to lift the overlying stable air significantly. The lifted layer can be quite thick, considering that large cities are wider than the scale thickness of the atmosphere (about 10 km). The downward buoyant force on a column of stable air lifted Δh above its surroundings is $\Delta h(\theta^{-1}\,\partial\theta/\partial z)\Delta p$, where θ is potential temperature and Δp is the hydrostatic pressure increment over the column's depth (weight per unit area). Upward force is supplied by the dynamic pressure change as CBL air slows over a city, $\rho\Delta U(U_0-\Delta U/2)$, where U_0 is the initial flow speed and ΔU is the amount it slows down. Equating these two forces, if $\theta^{-1}\,\partial\theta/\partial z \approx 10^{-5}\,\mathrm{m}^{-1}$ (typical for the troposphere) and $\Delta p \approx$ one half of sea level pressure, we find $\Delta h \approx (2.5\,\mathrm{m}^{-1}\,\mathrm{s}^2)\Delta U U_0$. For typical values of U_0 and ΔU in CBLs over cities (see Table I), this Δh is less than 10 m.

Thus, it seems likely that h enhancement is mostly growth caused by enhanced heating as air flows into an urban area, especially considering that maximum increases in h have been noted to occur near the time of maximum insolation (Spangler and Dirks, 1974). The amount of this heating depends a lot on the over-city fetch, Δx, and U, i.e., the time spent traveling over the area of enhanced heat flux, $\Delta x/U$. First-order models of h growth, with enthalpy conservation and θ constant with z in the mixing layer, give $h \sim (\int H^*\,\mathrm{d}t)^n$, where $\int H^*\,\mathrm{d}t$ reflects accumulated heating since sunrise and $n = 1/2$ if h grows through a layer of initially constant $\partial\theta/\partial z$ (however, because early morning stratification usually weakens with height, n tends to be somewhat larger). Letting t be the time since sunrise, $\bar{H}^* t = \int H^*\,\mathrm{d}t$ over rural areas, and ΔH^* be the enhancement of H^* within a city, we can write

$$1 + \Delta h/h \approx [1 + \Delta H^*(\Delta x/U)/(\bar{H}^* t)]^n\,, \tag{2}$$

where Δh is the increase in h within the city. For typical cases, $\Delta H^* \approx \bar{H}^*$. In the early morning, all the daytime travel of city-center air has been over enhanced H^* areas, so we set $\Delta x/U = t$; with $n = 1/2$, this yields $\Delta h/h \approx 0.4$. Later in the day, the percentage enhancement is much less, except with unusually large $\Delta x/U$. For example, $\Delta x = 18$ km and $U = 5\,\mathrm{m\,s}^{-1}$ gives only 1 hr of over-city travel.

5. Wind Speed Variations in Urban Areas

Diffusion as a function of distance may also be increased by a decrease in wind speed; this may result from larger surface roughness or from more vigorous vertical mixing of horizontal momentum caused by increased H^*. A typical urban value of roughness length, z_0, is 1 m, although it may be several times as large in areas with tall buildings. This contrasts with the $z_0 \sim 0.3$ m typical of suburbs and agricultural land with scattered trees. The effects on wind speed of changes in z_0 and H^* can be conveniently explored using a surface-layer similarity relationship for $\bar{u}/u^*$ based on an analysis of the Minnesota 1973 experimental data, where $\bar{u}$ is the mean wind speed at a given height and u^* is

friction velocity (Briggs, 1982):

$$k\bar{u}/u^* = \ln(z/z_0'), \qquad z_0' = z_0(1-4z/L)^{0.6}. \tag{3}$$

This relationship fit the wind profiles to $z \approx 0.2h$; because there is little wind shear in the mixing layer above 100 m in convective conditions, we can apply (3) at a height $z = 100$ m to approximate U with $\bar{u}$. Let us assume a certain initial speed, U_0, coming into the city over a roughness of 0.3 m, with $k = 0.4$ and $H^* = 25\ \mathrm{cm^2\ s^{-3}}$, a typical summer value over transpiring vegetation ($\bar{u}/u^*$ depends only weakly on H^*, so the value chosen is not critical). L is defined by $-u^{*3}/(kH^*)$. First, we can calculate the initial value of friction velocity, u_0^*, using (3). Then assume that H^* doubles and z_0 becomes 1 m in the city (this could as well be the case for air encountering a large forest). The mixing layer has much inertia and U does not significantly decrease until long after the lower part of the wind profile adjusts to the new surface. We can therefore calculate a new value of u^*, u_1^*, using (3) with $U_0 = \bar{u}$ at 100 m and the new surface parameters. Assuming that there is enough synoptic forcing to maintain the wind against a surface friction equal to u_0^{*2}, the initial frictional deceleration of the vertically-integrated mixing layer will be $h\,\mathrm{d}U/\mathrm{d}t = -\Delta u^{*2} = (u_0^{*2} - u_1^{*2})$. Finally, we can use (3) once again to calculate the new asymptotic windspeed, U_∞, that would occur if there were enough time or fetch to adjust back to $u^* = u_0^*$ over the new surface. Table I uses (3) to estimate these parameters, not with the precision of a PBL model, but with sufficient accuracy to estimate the magnitude of the urban effects on U due to increased surface drag.

First we note in Table I that the slowing of U is due mostly to increased surface roughness; only 26% or less of the slowing is due to doubled H^*, regardless of U_0. The initial increase in surface drag, Δu^{*2}, is approximately proportional to U_0, except for the $8\ \mathrm{m\ s^{-1}}$ case, which represents only slightly convective conditions ($-L \approx 240$ m). The initial rate of deceleration of the whole mixing layer, $-\Delta u^{*2}/h$, amounts to only $-1/4\ \mathrm{m\ s^{-1}}$ per 10 km of travel for all $U_0 \leq 4\ \mathrm{m\ s^{-1}}$, if $h = 1000$ m. Note that this will result in significant percentage increase in U^{-1} only for large $\Delta x/U$, exactly the case for significant h increase. The change in U also depends on h^{-1}.

In the last two lines of Table I we see that the distance for readjustment to U_∞ based on initial deceleration ranges from 24 to 73 km. This distance is closely approximated by $(U_0/u_0^*)^2 h/2$. Note that this is much larger than the distance required for readjustment of convective turbulence velocities and u^* over new surfaces in convective conditions, about $2(U/w^*)h$ (from Section 3). $(U_0/u_0^*)^2/2$ is typically about 50, while $2U/w^*$ is of the order of 5 in moderately convective conditions.

As wind speed decreases, the Coriolis acceleration lessens and the wind begins to turn slightly toward lower synoptic pressure. The turning rate in radians is only $\approx f\Delta U/U$, where ΔU is the change in wind speed and f is the Coriolis parameter, about $20^\circ\ \mathrm{hr^{-1}}$ at mid-latitudes. Considering the small values of $\Delta U/U$ on urban

scales, the turning of wind over a city is quite modest and the effect of lessened Coriolis on U is negligible (especially considering that it acts at nearly right angles to the initial wind direction). Another factor affecting U is the increased entrainment of momentum from above h if the growth rate of h is accelerated, as over a city. This factor may increase, decrease, or have negligible effect on U, depending on whether the mixing layer mean wind lags, opposes, or matches the wind just above h. This is difficult to generalize. This factor is most likely to be important in the early morning, when $h^{-1}\,\mathrm{d}\Delta h/\mathrm{d}t$ is the largest.

Local circulations induced by strong heat flux anomalies can also significantly affect U near the surface, especially when winds are light. Shreffler (1978) noted a systematic convergence of surface winds into the downtown area of St. Louis for daytime heat island cases, although the mean inflow speed for all 230 hours of data with $U < 3\,\mathrm{m\,s^{-1}}$ amounted to only about $0.4\,\mathrm{m\,s^{-1}}$, at a radius of 3–4 km. Wong and Dirks (1978) showed a more striking case study (21 July 1975) with about $1.5\,\mathrm{m\,s^{-1}}$ radial inflow near the St. Louis arch superimposed on a mean flow speed $\approx 3\,\mathrm{m\,s^{-1}}$; this resulted in $\bar{u} \approx 4.5\,\mathrm{m\,s^{-1}}$ near the west bank of the Mississippi, on the upwind side, and $\bar{u} \approx 1.5\,\mathrm{m\,s^{-1}}$ about 6–7 km to the SE, on the downwind side ($\bar{u}$ was measured at $z = 450\,\mathrm{m}$, about $h/4$). An order-of-magnitude analysis can be made for the inflow wind, v_r, created by the hydrostatic pressure difference generated by the increased buoyancy of an upflow over an area of radius R with ΔH^* excess buoyancy flux; the excess buoyancy in the upflow is ΔH^* divided by $\bar{w}$, the mean upflow velocity; the upflow volume flux over area πR^2 equals the radial inflow flux through a depth $h/2$, so $\bar{w}R = v_r h$. Driving the inflow with the hydrostatic pressure difference between $z = h/2$ (where zero inflow/outflow is assumed) and the middle of the inflow layer, at $z = h/4$, we can write $v_r^2/2 \approx (\Delta H^*/\bar{w})h/4 = \Delta H^* R/(4v_r)$, resulting in

$$v_r \approx (\Delta H^* R/2)^{1/2}\,. \tag{4}$$

For instance, $\Delta H^* = 25\,\mathrm{cm^2\,s^{-3}}$ acting over a radius $R = 3\,\mathrm{km}$ results in $v_r \approx 1.55\,\mathrm{m\,s^{-1}}$, reasonable for the case reported by Wong and Dirks (1978). Note the similarity between this expression for v_r and the definition of w^*. For a given city, the area of highest heat flux is relatively fixed, so R is nearly constant, while $\Delta H^{*1/3}$ undergoes only modest variations. Thus, inflow speeds do not vary much, but are likely to be noticeably present only when $U < 5\,\mathrm{m\,s^{-1}}$, when the heat column is not strongly sheared downwind.

7. Combined Urban Effects on Convective Diffusion Rates

Although there has been much study of the urban heat island and effects of cities on wind and turbulence, these physical effects are seldom introduced into diffusion modeling. Usually, urban diffusion measurements are analyzed using the conventional schemes categorizing diffusion discussed in Section 2, with crude empiricisms concerning σ_y and σ_z enhancement. However, in the framework of

TABLE I

Estimates of urban effects on U

	Assumed U_0 (m s^{-1})			
	1	2	4	8
U_∞ (m s^{-1})	0.43	1.1	2.6	5.7
Change due to H^*	26%	25%	24%	20%
u_0^* (m s^{-1})	0.14	0.22	0.36	0.61
u_1^* (m s^{-1})	0.21	0.31	0.48	0.79
Δu^{*2} (m^2 s^{-2})	0.024	0.046	0.100	0.25
For $h = 1000$ m, initial $-dU/dx$ (m s^{-1}/10 km)	0.24	0.23	0.25	0.32
$(U_\infty - U_0) \div dU/dx$ (km)	24	39	57	73
$(U_0/u_0^*)^2 h/2$ (km)	26	41	62	86

convective scaling, it is easy to make more quantitative estimates of urban effects on the growth of σ_y and σ_z. The degree of enhancement turns out to depend on city size and upwind U, among other things.

As noted by Weil and Brower (1982), convective scaling results suggest that a good daytime stability index is w^*/U. When $|h/L| \geq 10$, which corresponds roughly to Pasquill "C" or more unstable conditions, the spread of σ_y is proportional to w^*/U for a considerable distance (Briggs *et al.*, 1986; Hanna, 1986). The slope of descent of passive, elevated plumes is also proportional to $-w^*/U$, resulting in maximum ground concentrations near $x = 2\ z_s/(w^*/U)$, where z_s is source height (Briggs, 1985). In urban studies like that of McElroy and Pooler (1968), we see roughly linear growth in σ_z with x from a surface source before h is reached. For the most unstable classes, it is reasonable to assume that these σ_z are also proportional to (w^*/U). In fact, results from a recent field study show that the initial growth of both σ_y and σ_z is well fit by $0.6(w^*/U)x$ (Briggs *et al.*, 1986).

Thus, for any large surface anomaly, the question of most relevance for general convective diffusion effects is how much w^*/U is altered by the anomaly. As a primary example, let us assume that the surface anomaly is a city. How does a large urban area affect w^*/U? Three effects are fairly common: U may be decreased, h may be increased, and H^* may be increased. The net change in the diffusion rate, as characterized by w^*/U, depends on U_0 and on city size (a fact that is ignored in most urban diffusion modeling). We can estimate this change from the information in Sections 4–6.

For instance, for a large city with $U_0 = 2$ m s^{-1} and a 20 km fetch, h may increase by 20% and U may generally drop by 0.5 m s^{-1} (we ignore local surface convergence towards areas of very high heat flux; this may increase or decrease surface $\bar{u}$ by as much as 1.5 m s^{-1} on the upwind or downwind sides of the "hot spot"). Combining these changes with a doubling of the heat flux, the net factor

of increase in w^*/U is $(1.2 \cdot 2)^{1/3}(2/1.5) \approx 1.8$. We can expect a near doubling of σ_y and σ_z versus x compared with rural areas in these Pasquill "A–B" conditions. For this same fetch with $U_0 = 4\ \mathrm{m\ s^{-1}}$ (Pasquill "B–C" conditions), h may increase by only 10% and U is again reduced by about $0.5\ \mathrm{m\ s^{-1}}$, so the factor of increase in w^*/U is $(1.1 \cdot 2)^{1/3}(4/3.5) \approx 1.5$. Thus, diffusion enhancement is less in moderately convective conditions. At $U = 8\ \mathrm{m\ s^{-1}}$, the h increase may be only 5% and the wind speed drop at $\Delta x = 20\ \mathrm{km}$ may be $0.6\ \mathrm{m\ s^{-1}}$ (see Table I), yielding a factor increase in w^*/U of $(1.05 \cdot 2)^{1/3}(8/7.4) \approx 1.4$.

For the $U_0 = 8\ \mathrm{m\ s^{-1}}$ case, mechanical turbulence is dominant and conditions are only slightly convective. In this case, u^* significantly increases the diffusion rate. According to Weil and Brower (1982), it is enhanced by the factor $(1 + cu^{*2}/w^{*2})^{1/2}$, where $c \approx 8$ for σ_y and ≈ 5 for σ_z. For the example shown in Table I, this amounts to roughly 50% enhancement by mechanical turbulence of σ_y or σ_z at $U_0 = 8\ \mathrm{m\ s^{-1}}$, 20% at $U_0 = 4\ \mathrm{m\ s^{-1}}$, and 8% at $U_0 = 2\ \mathrm{m\ s^{-1}}$. Somewhat surprisingly, there is little change in these enhancement values after the air enters the city, because the increase in u^*, due mostly to increased z_0, is matched by an increase in w^*, due mostly to increased H^*. Because so many assumptions are involved about the character of surface change, this result may be seen as fortuitous, although representative of a typical urban situation.

The effects discussed in this section are not limited to urban areas. Ching *et al.* (1983b) reported aircraft measurements over rural areas in which the heat flux was 2.0 to 3.5 times as large over a forested area as over irrigated cropland. The roughness length may be on the order of 1 m over forest and 0.1 over cropland. If these surface inhomogeneities have scales on the same order as city inhomogeneities, they should have similar effects on turbulence and diffusion.

8. The Organizing Effect of Heat Flux Inhomogeneities

An interesting phenomenon seldom discussed outside the context of shoreline flows is the ability of heat flux inhomogeneities sometimes to "organize" convective flow structures. A striking example of this was seen in the St. Louis overflights discussed in Section 4 (Ching, 1985). The patterns of heat flux, H, and of the correlation coefficient for vertical velocity and temperature, R_{wT}, were very similar. This was especially well marked on the 18th, with its lower h, smaller U, and larger range of heat fluxes. R_{wT} reached a remarkably high value of 0.8 on the downwind side of the city, almost coinciding with the heat flux peak of $120\ \mathrm{W\ m^{-2}}$. The contour at $R_{wT} = 0.7$ almost matched the contour at $100\ \mathrm{W\ m^{-2}}$. Upwind, over the islands where H was $20\ \mathrm{W\ m^{-2}}$, R_{wT} plummeted to 0.2, a very low correlation; values of 0.4 to 0.6 were typical of the rural areas. It may be that the convective eddies over the river, having lost their driving force, also lost their organization, becoming more dissipative in nature. At the other extreme, $R_{wT} = 0.8$ indicates a high degree of organization of vertical motions by heat, or vice versa.

The latent heat flux, LE, and the correlation of vertical velocity with humidity fluctuations, R_{wq}, also mimicked each other in their patterns, but these were almost reversed from the H and R_{wT} patterns. On the 18th, R_{wq} reached a maximum at about 0.5 over the area of largest LE, 30 km upwind of the city. Here, the inverse Bowen ratio, LE/H, was approximately 5, and humidity fluctuations were large enough to help drive the turbulence. R_{wq} fell to an extremely low value, 0.1, at precisely the point where R_{wT} was maximal. This was also the driest area, with LE/$H \approx 0.8$. It appears that dry heat organized and drove the convection in this area, while humidity variations took a passive role. Latent heat may have come mostly from trees in this area, which were not the source of large heat fluxes. The areal distribution of H sources and LE sources may have been poorly correlated within the city.

An even more striking example of the organization of convection by heat flux inhomogeneities in St. Louis was shown in a paper by Kropfli and Kohn (1978). Metalicized chaff was released upwind of the city. It was observed by two doppler X-band radars. The analysis focused on the same horizontal planes, so that the radial velocities that each radar measured could be combined into a horizontal velocity vector. The mean velocity over the area was subtracted to get the eddy field, and the convergence field was calculated. This field was applied with the continuity equation integrated over z to infer vertical velocities. With this technique, Kropfli and Kohn discovered persistent, stationary horizontal rolls aligned with surface features of presumed heat flux anomalies. They presented average patterns observed for two periods: 70 min on 28 July 1975, with $U = 4$ m s^{-1} from 20° and $h = 1.55$ km, and 20 min on 4 August, with $U = 4.7$ m s^{-1} from 42° and $h = 1.3$ km. Both days showed a line of convergence near the surface, with a corresponding updraft, originating in the SW area of Granite City (Figure 2). This line maintained itself for the < 1-km trajectory across the Mississippi and continued over the heavily commercial and industrial part of St. Louis. Auer (1978) classified the southern portion of Granite City as light to moderate industrial, with < 5% vegetation.

The case of 4 August, which had very little mean wind shear, showed another interesting feature. There was a persistent line of divergence and associated downdraft in the NW part of the city, as shown in Figure 2. It was directly downwind of the Mosenthein and Gabaret Islands area, in which very small heat fluxes were observed the following year (Figure 1). Furthermore, the wind was nearly aligned with the long axis of this area. It seems likely that, just as the Granite City industrial area initiated upward motion, the river area of small H^* initiated the downdraft. The distance between these flow features, about 5 km, or 4 h, is about four times that usually observed between adjacent updrafts and downdrafts in convective rolls (LeMone, 1973). Therefore, it seems likely that the flow features really were tied to surface features, especially since they did not move. The circle through the mean wind speed arrow in Figure 2 shows the area of strongest downdraft, which was about 0.7 m s^{-1} at 600 m above the ground. A

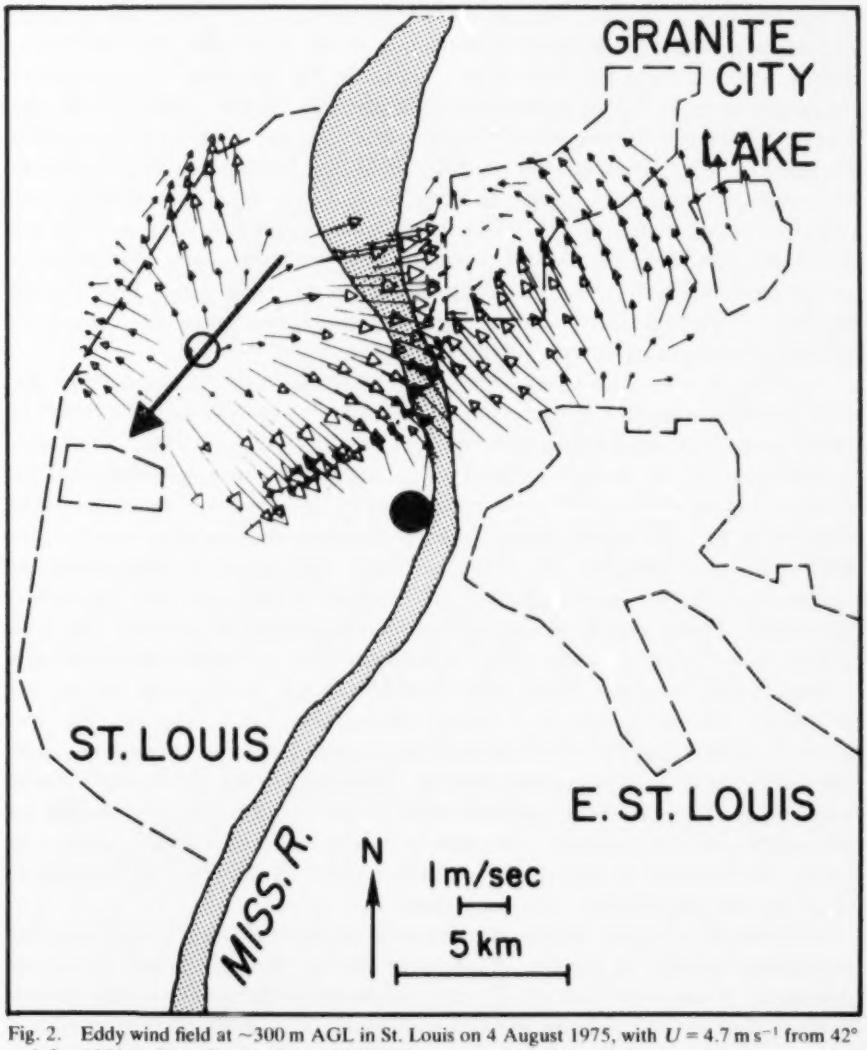

Fig. 2. Eddy wind field at ~300 m AGL in St. Louis on 4 August 1975, with $U = 4.7$ m s^{-1} from 42° and $h = 1300$ m (Kropfli and Kohn, 1978). Heavy arrow has been added to indicate mean wind velocity, and circle locates area of maximum downdraft velocity, about 0.7 m s^{-1}.

persistent downdraft of this magnitude can certainly be expected to intensify ground concentrations from elevated sources. This calls to mind an incident the author observed from an airplane during a 1975 field experiment. A plume from a 900 MW power plant (Bull Run), emitted from a 250-m stack, subsided directly over the waters of the Clinch River (TN), which is less than 1 km wide. This subsidence persisted for about an hour, on a very convective July day, when the wind direction happened to align with that section of the river.

The suggestion is made that convective scales are relevant in these cases. It is not the surface temperature difference that directly drives such circulations (this difference can be 20 to 25 °C, in the case of the Mississippi River at St. Louis). Rather it is the temperature difference or buoyancy imparted to the ascending air, which is most directly dependent on H^* differences. A more direct physical argument is that release of potential energy by vertical motion through h drives the kinetic energy. This is proportional to the buoyancy flux difference, ΔH^*, times h and divided by the velocity scale; thus, $\Delta w^* = (\Delta H^* h)^{1/3}$ is the velocity scale for organized updrafts and downdrafts. On the basis of the 1976 study, $\Delta H^* \sim 25\ \mathrm{cm^2\ s^{-3}}$, so for the day shown in Figure 2, $\Delta w^* \approx 1.5\ \mathrm{m\ s^{-1}}$. A typical magnitude of the measured downdraft was $0.6\ \mathrm{m\ s^{-1}}$, which is 0.4 Δw^* in this case. This agrees with other estimates of downdraft magnitude (e.g., Briggs 1975) and is about the same magnitude as the convective turbulence downdrafts that cause higher ground concentrations from elevated sources than predicted by Gaussian models. However, the latter downdrafts are advective and occur intermittantly at a source point only about 60% of the time, whereas the rolls observed by Kropfli and Kohn (1978) were persistent, stationary features. The effect of a persistent downdraft on maximum ground concentrations from a point source might be modeled simply by subtracting $0.4(\Delta w^*/U)x$ from the effective source height, in addition to other model features.

Both fetch and wind alignment are important. It takes several nondimensional X, or $x = (U/w^*)h$, of travel over a heat flux anomaly for the full effect of Δw^* to take place. A small industrial area or lake may not intiate these flows except on low wind speed, convective days with relatively small h. Rivers, even the Mississippi, have little effect when a moderate or strong wind crosses them at a large angle. However, even a small river can cause persistent subsidence when the wind aligns with a straight section of the river, if its width is at least several tenths of h.

9. Summary and Conclusions

Traditional methods of classifying diffusion rates by stability categories fail to account for the effects of the mixing depth, h, which overshadows the effects of all other meteorological variables except for mean wind speed, U, in convective conditions (defined here by $h > -10L$). Furthermore, crude jumps in stability category or multiplying factors for "urban" diffusion rates fail to allow for the continuum of σ_y or σ_z enhancement that actually exists, depending on city size and U, among other factors. It is suggested that extensions of convective scaling provide the best framework for estimates of urban effects on diffusion rates and other diffusion effects due to inhomogeneous surfaces, during convective conditions. The best estimates of diffusion speeds are based on w^*, the convective scaling velocity, or on Δw^*, defined on the basis of differential heat fluxes over contrasting surfaces.

The rates of adjustement of turbulence and wind to new underlying surfaces were estimated. Wind entering a city or a forest slows down but adjusts slowly, usually never reaching a new equilibrium speed, because the required fetch is on the order of $(U_0/u_0^*)^2 h/2$. In contrast to this, the fetch required for convective turbulence adjustments to changes in heat flux is characteristically $\sim 2(U/w^*)h$, normally just a few km. Turbulent variances measured at 150 m over the greater St. Louis area scaled to the local w^* within 10% of the expected values over about 80% of the area. The greatest departures were in areas of sharp heat flux contrasts, over and near the Mississippi River. The response of measured turbulence to the heat flux inhomogeneities may be crudely explained in terms of the dimensionless fetch, convectively scaled as above.

Urban influences on heat flux, h, and U were estimated. Sensible heat flux is typically double rural values in urban areas, but may be twice this or may be less than rural values in a desert city; it largely depends on how much solar energy is expended on evaporation and vegetative transpiration. "Doming" of h is frequently observed over large cities. This is not likely due to convergence caused by wind speed reduction, as the kinetic energy reduction in convective conditions is too small to account for the observed degree of lifting (sometimes 400 m). The h growth seems likely to be caused by the greater rate of accumulation of heat flux as the mixing layer traverses a city. The ratio of urban to rural h thus depends on the ratio of heat fluxes, and is equally influenced by the traverse time, $\Delta x/U$. It can be estimated using Equation (2). It ranges from 1.0 to about 1.5, the latter being possible only for a large city during very low U. Wind speed may be decreased, mostly by increased roughness in a city, but the drop induced in U is only of the order of $1/4\ \mathrm{m\,s^{-1}}$ per 10 km of fetch, when $h \approx 1000$ m (the deceleration is $\sim h^{-1}$). During low U conditions, there may be noticeable surface convergence towards a concentrated heat island like a city center, with radial velocities of the order of $1.5\ \mathrm{m\,s^{-1}}$.

The combined effects of heat flux and roughness length inhomogeneities on convective diffusion were considered. Using w^*/U as a convenient, direct index of convective diffusion rates, we explored its increase in urbanized areas due to enhanced heat fluxes, somewhat larger mixing depths, and reduced U caused mostly by larger surface roughness. At low wind speeds in large cities, w^*/U approximately doubles. It changes less at larger wind speeds and in small cities. The convective character of the turbulence, as indicated by ratios like w^*/u^*, does not necessarily change much as air enters a city, even though the turbulence intensity is greater; w^* and u^* can increase about equally. Large forests surrounded by cropland can have similar effects on turbulence and diffusion, such as enhancement of w^*/U.

Finally, cases of highly organized convective motions near urban heat flux maxima and minima were noted, using several studies of the St. Louis area. High heat flux areas had unusually large correlations between turbulent vertical velocities and temperature fluctuations. Also discussed were cases of persistent,

stationary convective rolls that appear to have been caused by heat flux anomalies. A line of surface convergence and updraft was probably initiated by an industrialized area, and a line of surface divergence and downdraft probably originated over a 4-km wide area of river and low wooded islands, surrounded by city. The persistent downdraft had velocities of about $0.6\,m\,s^{-1}$, enough to increase surface concentrations from elevated sources considerably. It was suggested that convective scaling is relevant to these motions; the downdraft magnitude was about $0.4\ \Delta w^*$. Such circulations are likely to occur near any river wider than $\sim h/4$ on convective days if the wind direction is approximately aligned with the river.

Acknowledgements

The author thanks J. M. Godowitch and J. K. S. Ching of the Meteorology and Assessment Division, ASRL, USEPA for discussions and analyses of urban turbulence measurements in St. Louis, J. K. S. Ching for the use of Figure 1, and R. A. Kropfli for the base figure modified here in Figure 2.

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